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Crustal Structure and Surface Wave Dispersion Part II
Solomon Islands Earthquake of 29 July 1950

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(Columbia University)

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by

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ABSTRACT

Rayleigh waves from the Solomon Islands earthquake of 29 July 1950 recorded at Honolulu, Berkeley, Tucson and Palisades are analyzed. Both the direct waves and those propagated through the Antipodes were observed for all stations except Honolulu. Application of a correction for land travel results in a dispersion curve for the oceanic portion of the path. It is found that the observed dispersion could be accounted for by propagation through a layer of water 5.57 km thick overlying simatic rocks having shear velocity 4.56 km/sec and density 3.0 gm/cc. Basement structure in the Pacific, Indian, South Atlantic and North Atlantic oceans is identical within the limits of accuracy of the method.

The sinusoidal nature and duration of the coda is explained by the effect of the oceans on the propagation of Rayleigh waves.

The results are compatible with seismic refraction measurements in the Atlantic and Pacific Oceans.

I. Introduction

In a previous paper[1] the authors reinterpreted published readings of travel times of surface waves of various periods in the light of their theoretical treatment of the effect of the ocean water upon Rayleigh wave propagation[2]. The data of Wilson and Baykal[3] and Bullen and DeLisle[4,5] were used. It was concluded that the observed Rayleigh wave dispersion could be explained by a layer of unconsolidated sediments about 0.7-1.3 km thick, which was considered as an addition to the water depth, underlain by a considerable thickness of rock in which the velocities of shear and compressional waves were about 4.45 and 7.71 km/sec respectively, assuming Poisson's constant to be 0.25. In Part I[2] it was pointed out that the very small dispersion found by Wilson[6] for sub-oceanic Love waves indicated that these waves were propagated in a layer

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1 Maurice Ewing and Frank Press, "Crustal Structure and Surface Wave Dispersion", Bull. Seism. Soc. Amer., 40: 271-280, 1950.

2 Frank Press, Maurice Ewing, and Ivan Tolstoy, "The Airy Phase of Shallow Focus Submarine Earthquakes", Bull. Seism. Soc. Amer., 40: 111-148, 1950.

3 J.T. Wilson and Orhan Baykal, "Crustal Structure of the North Atlantic Basin as Determined from Rayleigh Wave Dispersion", Bull. Seism. Soc. Amer., 38: 41-53, 1948.

4 K.E. Bullen, "On Rayleigh Waves Across the Pacific Ocean", Mon. Nat. Roy. Astron. Soc., Geophys. Suppl. 4: 579-582, 1939.

5 J.F. DeLisle, "On Dispersion of Rayleigh Waves from the North Pacific Earthquake of November 10, 1938", Bull. Seism. Soc. Amer., 31: 303-307, 1941.

6 J.T. Wilson, "The Love Waves of the South Atlantic Earthquakes of August 38, 1933", Bull. Seism. Soc. Amer., 30: 273-301, 1940.

whose elastic properties differed so slightly from those of the substratum that the rocks could be treated as a unit for Rayleigh wave propagation, permitting use of theoretical curves already available. Recent seismic refraction measurements in the Atlantic[7] and in the Pacific[8] give a sediment thickness of the order used in the dispersion study, and give a basement layer about 5 km thick with compressional wave velocity about 6.8 km/sec overlying rock having a compressional wave velocity of 8.1 km/sec. The velocity 7.71 km/sec deduced from the dispersion study is an acceptable average for the basement layers. Calculations of Rayleigh wave propagation, taking account of the basement layering, are lengthy but these are now under way.

The object of the present paper is to study earthquake surface waves in more detail, and to produce additional evidence that the observed dispersion of Rayleigh waves in oceanic paths is controlled by the water layer instead of by a sialic upper layer in the basement rocks. The crucial point in the test is the group velocity of the Rayleigh waves of periods about 15 to 16 seconds, which should approach the speed of sound in water in one case, but a speed more than twice as great in the other. The authors consider that the study has confirmed the conclusions about crustal structure beneath ocean basins which was reached in Part I.

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[7] C. B. Officer Jr., Maurice Ewing, Paul C. Wuenschel, "Seismic Refraction Measurements in the Atlantic Basin - Part II", presented at Los Angeles meeting Seism. Soc. Amer., 1951.

[8] Russel W. Raitt, "Seismic Refraction Studies of the Pacific Ocean Basin", presented at Los Angeles meeting Seism. Soc. Amer., 1951.

It has also produced an explanation for the duration of the coda and for the sinusoidal character and periods of the waves which compose it, for all seismograms studied.

For the present study the authors have selected the earthquake of 29 July 1950 (23-49-08; 6.8°S 155.1°E, depth 75 ± km J.S.A., Magn. 7 Pas.) and the seismograms from Honolulu, Berkeley, Tucson, and Palisades stations which lie near to a single great circle through the epicenter, as shown in Figure 1. Corrections for the continental part of each path are made by use of Wilson and Baykal's continental dispersion curve[3] as plotted in Figure 5, and the velocities for the oceanic part of each path are computed for a number of periods. These velocities agree with the theoretical dispersion curve for essentially all surface waves on the seismograms, including those which have come through the Antipodes, for all stations except Honolulu, where only the direct waves were strong enough to be recorded at the sensitivity used.

II. Experimental Methods and Results

Until recent years the method of reading dispersion data from seismograms was to take the period and arrival time for the dominant or the first surface wave as recorded at a large number of stations, obtaining but a few points on the curve from each station. Wilson and Baykal[3] read the period and arrival time of a succession of waves in the early part of the Rayleigh wave train. As these authors state, this method has many advantages. In the present work periods and arrival times are read for the entire Rayleigh wave train. This method is of great value in dispersion studies and is essential for studies of the spectral distribution of energy in Rayleigh waves. An example of the method is shown in Figure 2 made from readings of a Palisades vertical seismogram (To=12 sec Tg=15 sec). Arrival times for crests are plotted against crest number. Tangents are drawn to the smooth curve through the points. From these period and arrival time are read and the group velocities are calculated by obtaining the travel time for waves of each period and applying the correction for continental travel described above.

In addition to three component Columbia seismograms (To=12 sec Tg=15 sec), three component Benioff seismograms (To=1 sec Tg=75 sec) are available at Palisades. Arrival times for troughs and crests of waves arriving over a period of several minutes are plotted in Figure 3 for the Benioff instruments and for a Columbia vertical. Although the two types of seismograph have quite different phase response, it is seen that period-travel time data from two instruments will be in agreement, hence it is neither necessary to integrate the seismogram for true ground motion nor to attempt to eliminate instrumental phase shift by other methods in order to obtain dispersion curves free from distortion within the precision required for this study. Figure 4 is a tracing on a common time scale of a portion of the Rayleigh wave train as registered at Palisades on the three component Columbia seismographs. From either Figure 3 or 4 it can be verified that the oscillations consist of Rayleigh type waves arriving at Palisades from the NW quadrant.

For the latter part of the seismogram the variation of period over an interval of several cycles is usually very small and the dispersion curve can be obtained more quickly and without loss of accuracy by measuring the average period of 5 to 10 waves selected from well-formed groups.

The epicentral distances for all stations are given in Table 1. The paths are computed along great circles, with no allowance for refraction. Since the Honolulu, Berkeley and Tucson direct paths could involve only very small corrections for continental paths or for refraction effects, the agreement of the dispersion data from these paths with that from the others is evidence that no serious error has been introduced. The land part of each path has been taken as the total distance between 1000 fathom curves for continental areas lying on the great circles in question.

Toward the end of the coda of the seismograms of all stations except Honolulu, the regular decrease in period which had prevailed earlier was interrupted by a train of longer period waves whose orbital motion identified them as Rayleigh waves from the direction of the Antipodes. The period and travel time were obtained and the group velocity was calculated, using the distance through the Antipodes as listed in Table 1 and correcting for the land portion of the path.

The observed group velocity values for both paths for each station are plotted in Figure 5. These are to be compared with the theoretical curve computed for the case $\beta_2 = 3\alpha_1 = 4.56$ km/sec, $\alpha_1 = 1.52$ km/sec, $\alpha_2 = \sqrt{3}\beta_2 = 7.90$ km/sec, $\rho_2 = 3\rho_1$, $H = 5.57$ km. as indicated in the figure. Here H is taken as the combined depth of water plus sediment.

Table 1

	Direct Path				Path through Antipodes		
Station	Total	Land km	Ocean km	Azimuth	Total	Land km	Ocean km
Honolulu	5970	0	5970	244°			
Berkeley	9810	0	9810	261°	30,190	9400	20,790
Tucson	10,780	560	10,220	267°	29,220	9300	19,920
Palisades	13,850	4000	9850	295°	26,150	4550	21,600

III. Discussion

1. A striking feature of these seismograms is that the Rayleigh wave period continuously decreases until the corresponding group velocity is less than the speed of sound in water. Figure 5 shows that this period change is related to group velocity as predicted by the theory of propagation in a liquid layer in contact with a homogeneous solid.

2. Figure 1 shows that the direct paths cross only the Pacific Ocean, while the paths through the Antipodes cross the Indian, South Atlantic and North Atlantic Oceans. The agreement between observed group velocities indicates that the basement structure in all the

ocean basins traversed is identical within the limits of accuracy of the method.

Numerical calculations on the propagation of Rayleigh waves in three layers are not yet complete. A good fit of observations using two-layer theory with water and sediment as the layer gives 4.56 km/sec and 7.90 km/sec, respectively, for shear and compressional waves in the substratum. The Rayleigh waves on which this determination is based range in length from about 30 to 150 km. They are affected by the elastic properties of basement rocks to depths of the order of their lengths, the rocks nearer the surface having the greater influence for the shorter wave lengths. We consider the velocity of 7.9 km/sec, which fits the dispersion data on the assumption of a homogeneous basement, to be compatible with the refraction results which indicate a 5 km thick layer with longitudinal wave velocity of 6.8 km/sec underlain by a thick layer with velocity about 8.1 km/sec. The 8.1 km/sec horizon is identified with the Mohorovicic discontinuity which is at depths of 30 to 40 km beneath the continents.

3. A long standing problem in seismology has been the sinusoidal nature of the coda and its duration, which indicated group velocities as low as 1.5 km/sec, which was less than half of any known wave controlled by solid layers. The dispersion theory applied in this paper accounts for the periods and arrival times of all coda waves observed on the seismograms. In the theory[2] for Rayleigh waves generated by an impulsive point source and propagated along an oceanic path it is shown that the disturbance at large ranges consists of a train of sinusoidal

waves whose period and travel time can be obtained from theoretical group velocity curves such as that presented in Figure 5. Stated otherwise, the impulse delivered at the focus may be considered as the summation of a family of long sinusoidal wave trains whose frequencies cover the whole spectrum, and whose phases and amplitudes are chosen so the sum is initially zero except in the focal region. In a dispersive medium these trains travel at velocities appropriate to their frequencies, hence the shape of the impulse changes with time. At a considerable distance from the source the summation of the family of wave trains will result in a long train of sinusoidal waves of gradually changing period, waves of each period having been propagated at the group velocity given by the dispersion curve.

4. The observed Love wave dispersion for the oceanic part of the paths can be accounted for by the 5 km layer discussed above. No detailed study of the Love waves in these seismograms was made. It was established, in full agreement with the work of Wilson[6] that both the Love wave amplitudes and the dispersion are small in contrast to their values for continental paths.

5. An objection has sometimes been raised against the idea that long trains of Rayleigh waves, such as those studied here, could occur solely through the effects of dispersion. The view has often been expressed that a long succession of sinusoidal waves of almost constant period was due to some resonant phenomenon in general. Likewise it has been maintained that the absence of long period surface waves at the smaller epicentral distances precludes the possibility that dispersion can be responsible for the appearance of the train at

great distances. The portion of the coda in which the period seems to remain constant results from the steep portion of the dispersion curve where a large change in group velocity occurs for a very small change in period. Our experience in the study of many seismograms chosen to avoid serious complications from refraction at continental margins is that careful reading will always show a continual decrease in period until the arrival of the waves from the direction of the Antipodes.

The fact that travel over a considerable distance is necessary before the long waves are visible at the head of the train depends on the fact that the slope of the group velocity curve is very small for long periods, approaching zero as the period approaches infinity. The wave train to be expected at a given distance may be deduced from the dispersion curve by a procedure which is exactly the reverse of that used above to discover the dispersion curve from measurements on a seismogram. Due to the maximum of group velocity which occurs at infinite period there will exist a first wave in the train which will travel at the velocity of Rayleigh waves in the bottom, $0.92\beta_2$. Shorter periods will lag behind, by an amount proportional to the distance and determined by their lower group velocities. A curve showing the period of waves to be expected at any time subsequent to $\Delta/0.92\beta_2$ may be constructed by simply converting the ordinates in the dispersion curve of Figure 5 into travel times for the chosen distance. If this is done it will be seen readily that at short distances the period drops from infinity to a very low value before the first cycle of the wave train can be completed. At larger distances many cycles would be completed for the same spread of period, and the first cycle of the train would have a correspondingly longer apparent period. If we consider the actual earth instead of the ideal two layer structure for which our group velocity curve was calculated, waves of very long period will show an additional increase in group velocity due to the effect of deeper layers, but this does not affect our argument for the range of periods under discussion in the present paper.

Figure 1. Great circle path from epicenter to Berkeley

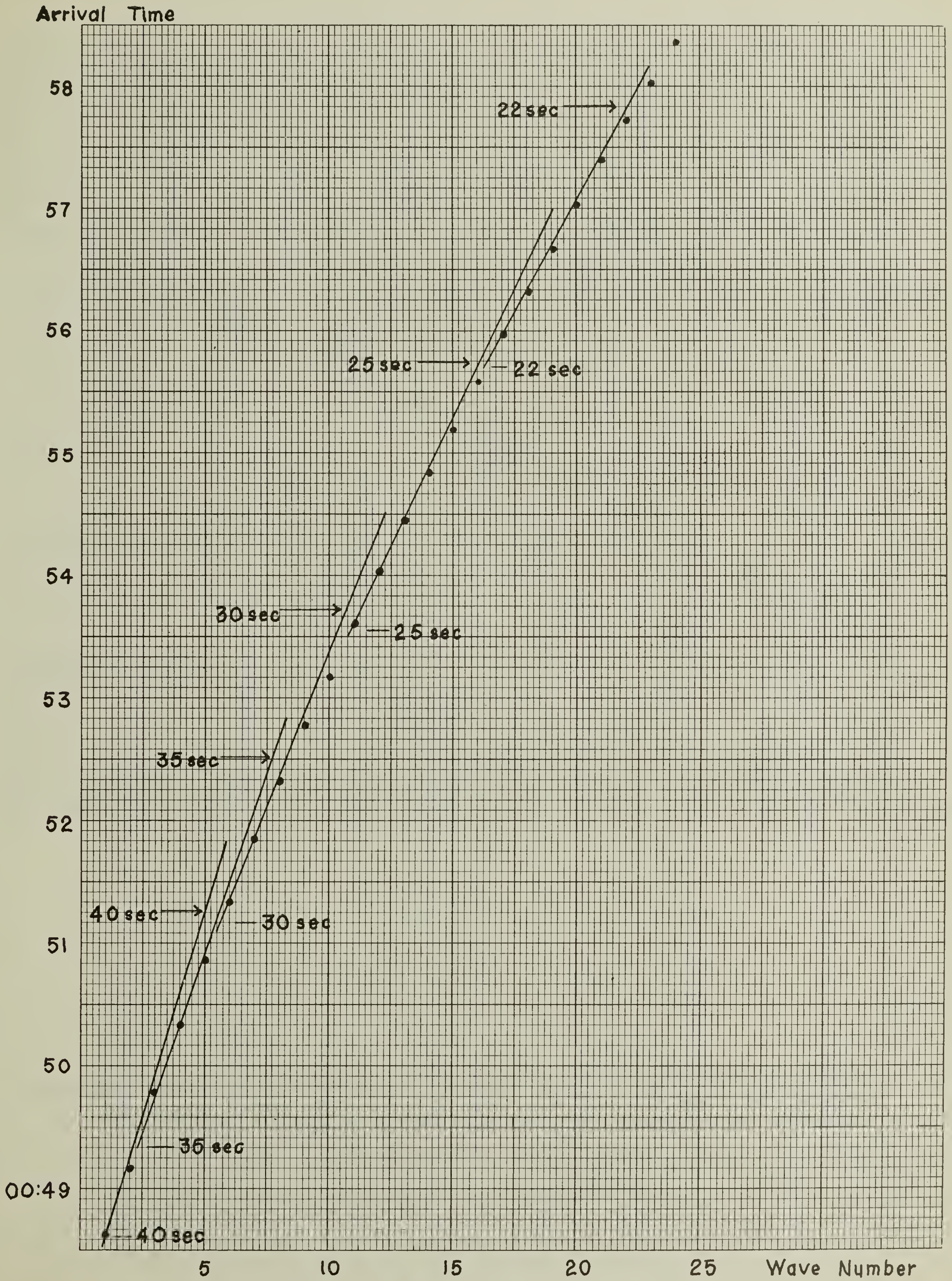

Figure 2. Example of method of obtaining arrival time of waves of different periods in early part of seismogram.

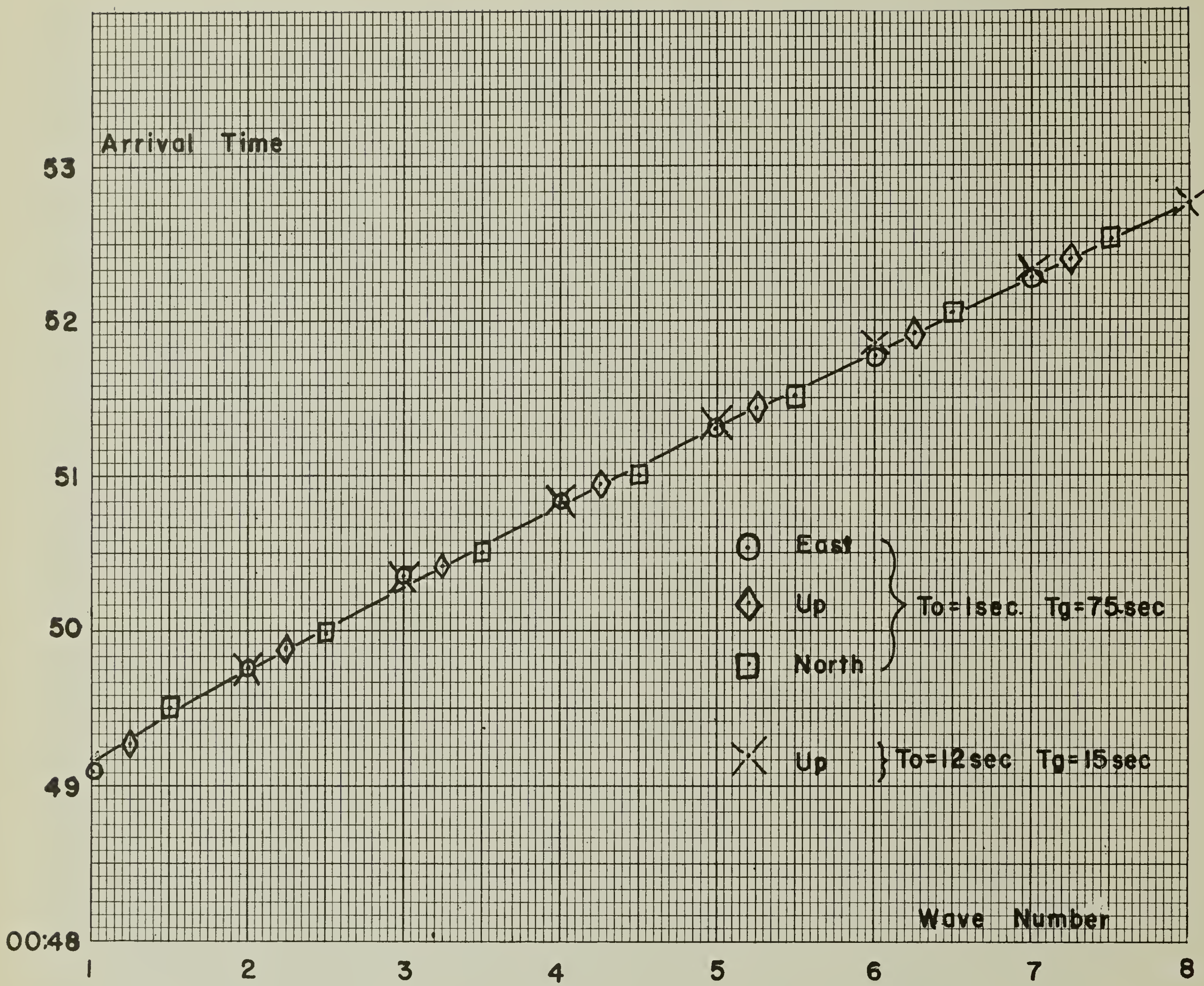

Figure 3. Comparison of readings at Palisades from three component Benioff Seismographs (T_o = 1 sec, T_g = 75 sec) and Z component Columbia Seismograph (T_o = 12 sec, T_g = 15 sec).

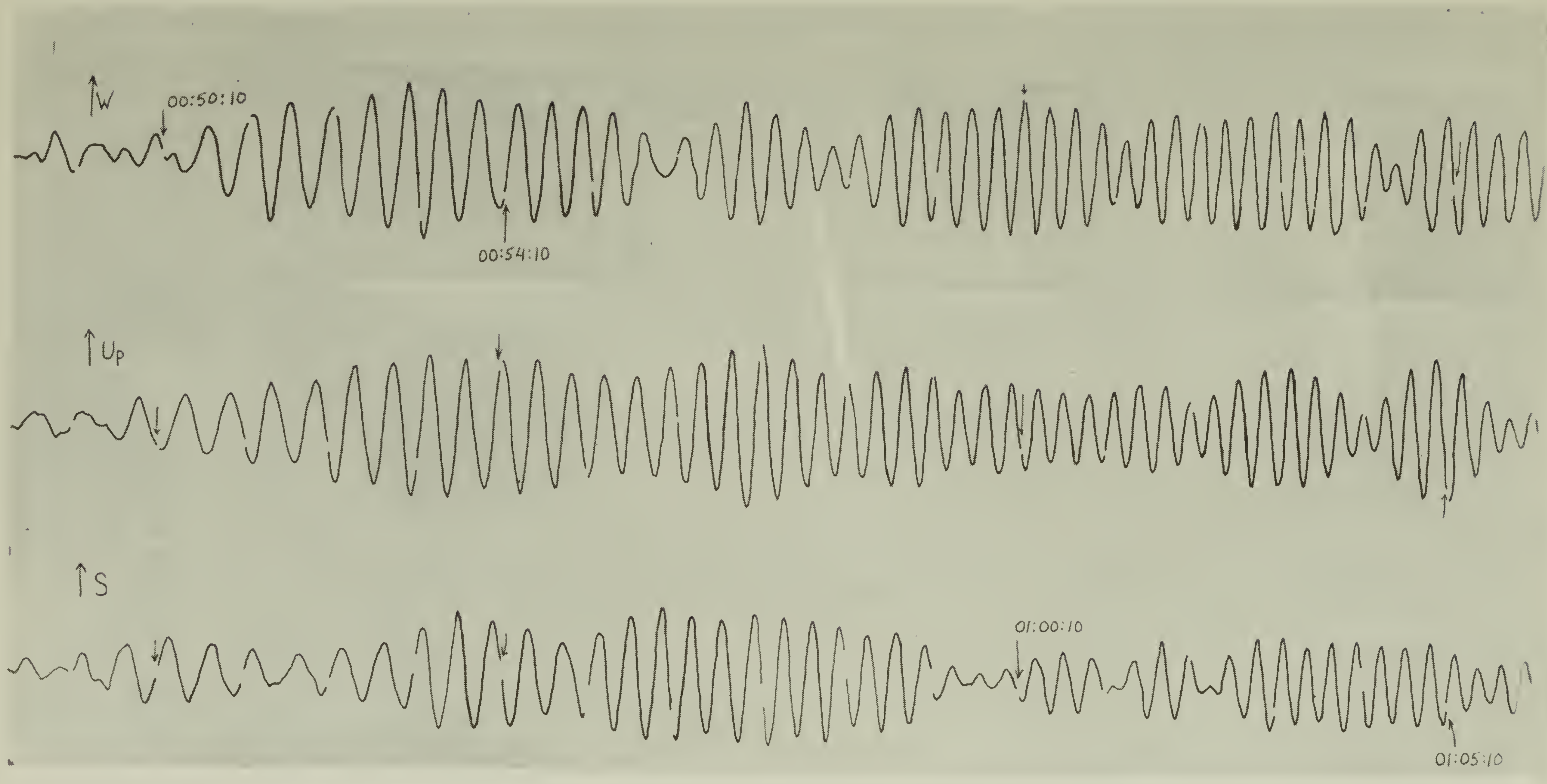

Figure 4A. Tracings of three component seismograms (T_o = 12 sec, T_g = 15 sec) for Palisades.

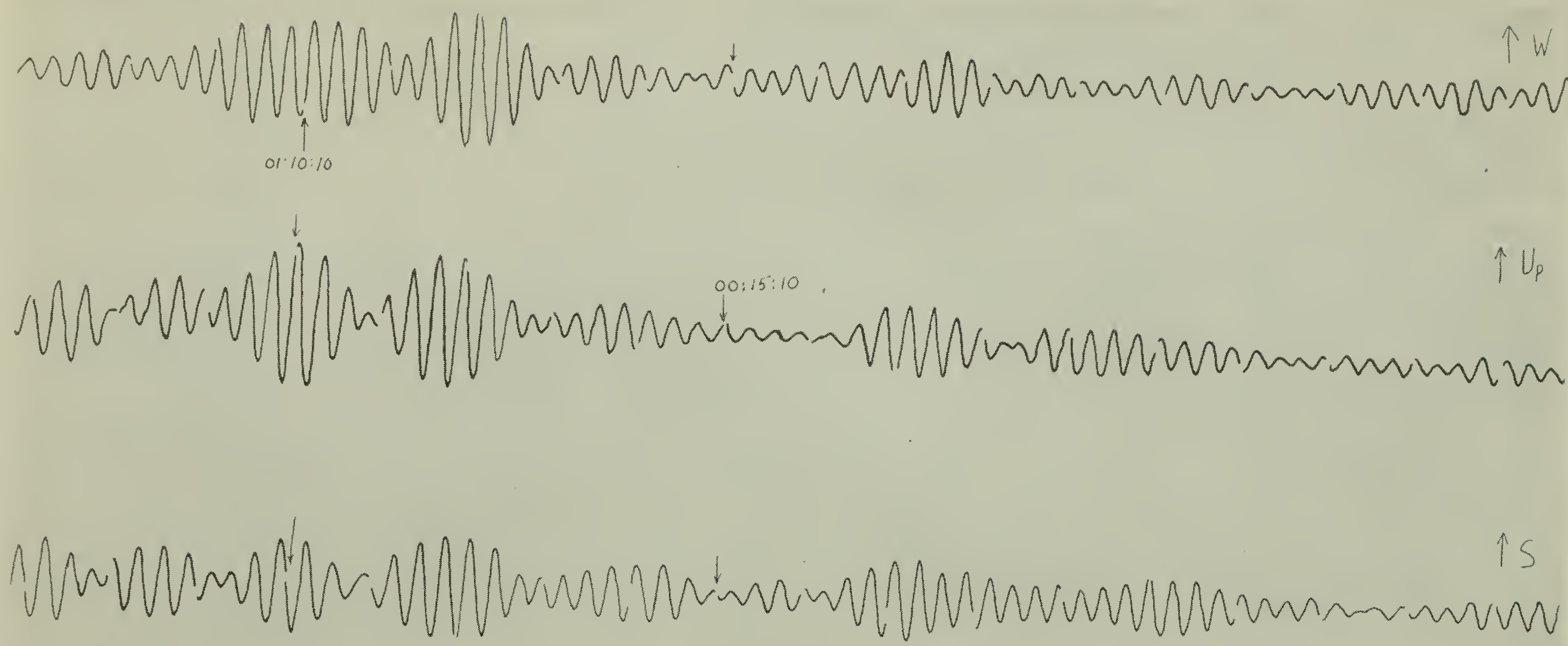

Figure 4B. Continuation of Figure 4A.

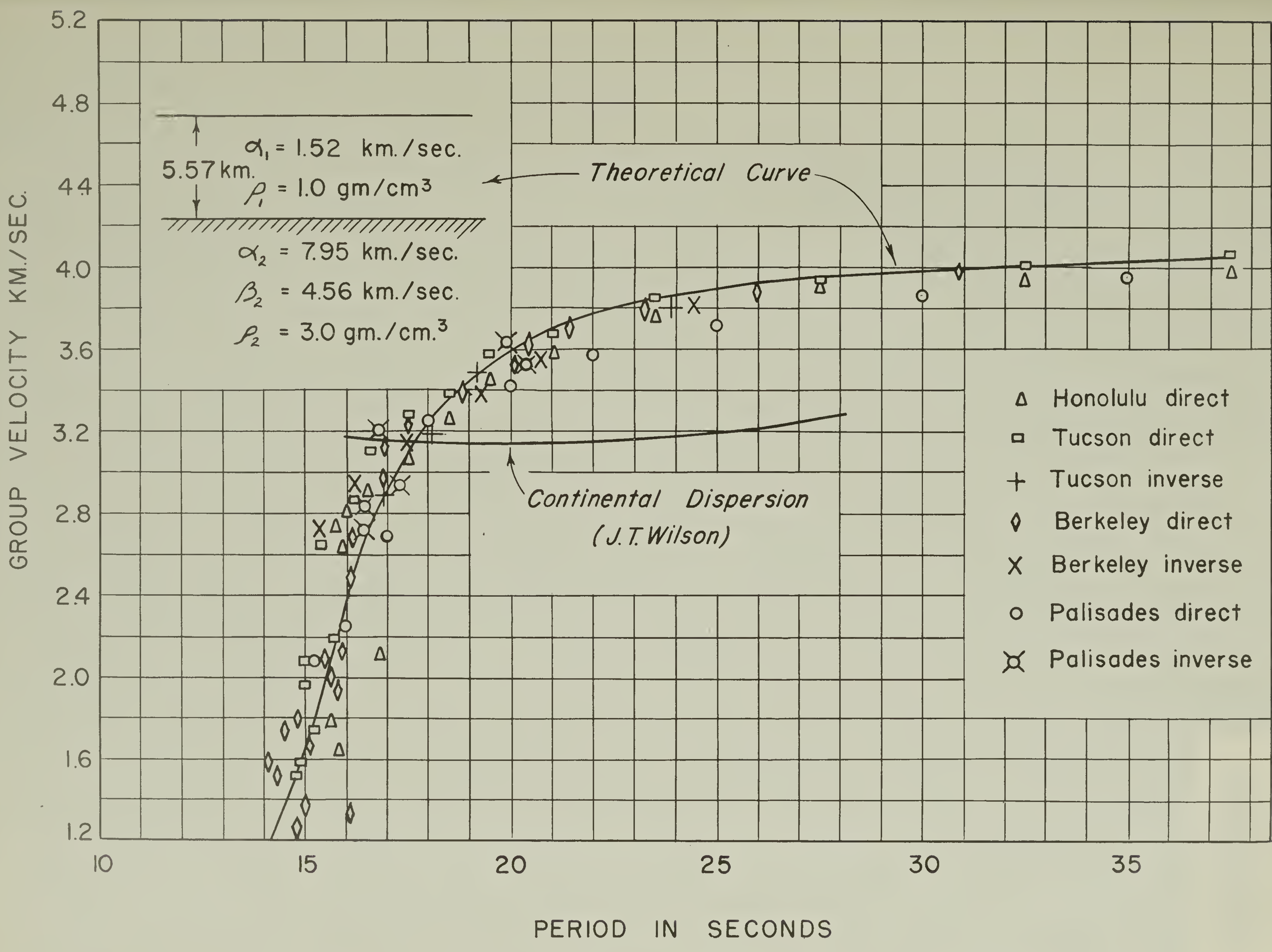

Figure 5. Observed and theoretical dispersion curves for oceanic paths to Honolulu, Berkeley, Tucson, and Palisades.